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**ABSTRACTS FOR THE PLANETARY GEOLOGY FIELD
CONFERENCE ON THE SNAKE RIVER PLAIN, IDAHO**

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October 1977



Abstracts
for the
Planetary Geology Field Conference
on the
Snake River Plain, Idaho

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1

STRUCTURAL AND VOLCANIC CHARACTERISTICS OF THE EASTERN SNAKE RIVER PLAIN, IDAHO

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Detailed reconnaissance mapping on the eastern Snake River Plain, an area of about 19,000 km² located east of 114°15' W. Long and north of 43° N. Lat, has been completed at a scale of 1:24,000. These studies show that this continental basalt province is characterized by distinctive structural and volcanic features.

The major structural features of the area studied in the eastern Snake River Plain are rift zones, which are defined by rectilinear distribution of structural and volcanic features such as graben, extensional faults, cinder cones, eruptive fissures, shield volcanoes whose vents are elongated over inferred rifts, and non-eruptive open fissures. These features are typically arranged along the rift zones in the order listed above from the north margin of the Plain to the Plain axis. Similar relationships are not clearly recognized near the south margin of the eastern Plain, probably owing to erosion and deposition of sediments by the Snake River. The largest number of rift zones trend NW-SE and N-S, normal to the long axis of the Plain. These include the Great Rift, Arco, Spencer-High Point, Menan, Hells Half Acre-Lava Ridge, and Rock Corral Butte rift zones. A few rift zones with NE-SW and E-W trends are present in the study area, but they are short and relatively discontinuous. Some rift zones are narrow and have had volcanism centered on them for only short periods of time (e.g., Menan); others are narrow zones in which volcanism has been sporadic over long periods of time (e.g., Hells Half Acre-Lava Ridge); and others are broad zones with indistinct boundaries where volcanism has also been sporadic (e.g., Great Rift and Arco). Rhyolite domes (Big Southern Butte, East Butte, and Middle Butte?) and a ferro-latitude volcano (Cedar Butte) all occur at the intersections of NW-SE rift zones with rift zones of different trends.

Nearly all of the Pleistocene and Holocene volcanoes in the study area represent rift-controlled eruptions; their distribution is not "random" as has been stated by some workers. Although varied volcanic landforms occur, shield volcanoes located near the center of the Plain, covering areas between 200 and 500 km² and having volumes on the order of 50 km³, make up the bulk of the volcanic pile along the axis of the eastern Snake River Plain. In contrast, the margins of the eastern Plain comprise volcanic products chiefly from cinder cones, spatter ramparts along eruptive fissures, and smaller shield volcanoes, and interlayered sediments.

Rift zones that intersect the Plain margins (those having NW-SE and N-S trends) are aligned with range-front faults and older structures in mountains adjacent to the Plain. The fault zones beyond the Plain margins and their rift-zone extensions onto the Plain define continuous, nearly rectilinear structural systems, all parts of which are related to the regional NE-SW extension that characterized the late Tertiary-Quaternary stress field of the northeastern Great Basin. Dominantly dip-slip or strike-slip fault movement has been documented in the mountains north and south of the Plain, whereas extensional fissures are the only

obvious expression of strain at the surface near the axis of the eastern Plain. The continuation of these structural systems across the margins of the eastern Plain, the absence of identifiable boundary faults for the eastern Plain, and the lack of clear geophysical evidence for such faults suggest that the eastern Plain is not a graben decoupled from the surrounding mountains, but is rather a downwarp similar to that originally proposed by Kirkham (1931).

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Computer Classification of Pahoehoe flows in the Craters of the Moon Volcanic Field, Idaho Using Landsat data.

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M.J. Abrams, Jet Propulsion Laboratory, California Institute of Technology, Pasadena, California, 91103.

The major lava flows of the Craters of the Moon (COM), Idaho area were computer classified on the basis of their Landsat MSS spectral radiances. The radiance values (expressed as digital numbers, or DN) of the pahoehoe flows of the COM volcanic field (Figures 1 and 2) show a close correlation with flow age: the flows with higher DN values generally are older than those with lower DN values. Also, the older flows exhibit less change in DN values from Landsat band 4 through band 7. Both of these characteristics can be explained by the masking effects of increasing degrees of weathering and cover by vegetation and sediment with time on the original bluish flow surfaces. (Lefebvre, 1975).

Figure 1 shows, from oldest to youngest, the progressive decrease in DN and increase in slope from MSS band 4 through band 7. The Pleistocene Snake River Plain flows (SRP) have the highest DN and least change between the four Landsat bands.

The oldest COM, intermediate COM, and youngest COM units of Figure 1 were further subdivided into 11 classes of individual flows on the basis of field relationships (Figure 2). The names given to these flows are informal. The Kimama, Fingers Butte and Sunset flows are the flows used in the oldest COM pahoehoe flow unit of Figure 1. The Sunset flow has been radiocarbon dated at $11,300 \pm$ y.b.p. (Lefebvre and Abrams, 1976). The Carey and Minidoka flows were the two flows used in the intermediate COM pahoehoe unit of Figure 1. The Light and Dark Blue Dragon flows make up the youngest COM pahoehoe unit, and are dated at approximately 2000 y.b.p. by a Bullard and Rylander (1970, 1971). The Grassy, Pronghorn, and Laidlaw Lake flows were not included in the four class classification (Figure 1).

On the basis of field data, exceptions to the correlation between age and radiance values are 1) Carey is older than Grassy; 2) Laidlaw Lake is older than Grassy; 3) Pronghorn is older than Minidoka. In the field the surfaces of Laidlaw Lake and Pronghorn have greenish glassy crusts rather than bluish crusts common to most of the pahoehoe flows in the COM area. These crustal differences may account for the largest discrepancies (nos. 2 and 3 above) in the correlation between age and DN values. The discrepancy between the Carey and Grassy flows (no. 1 above) is not great and may indicate that the ages of these flows are similar.

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CRATERS OF THE MOON VOLCANIC FIELD, IDAHO

11 CLASS CLASSIFICATION

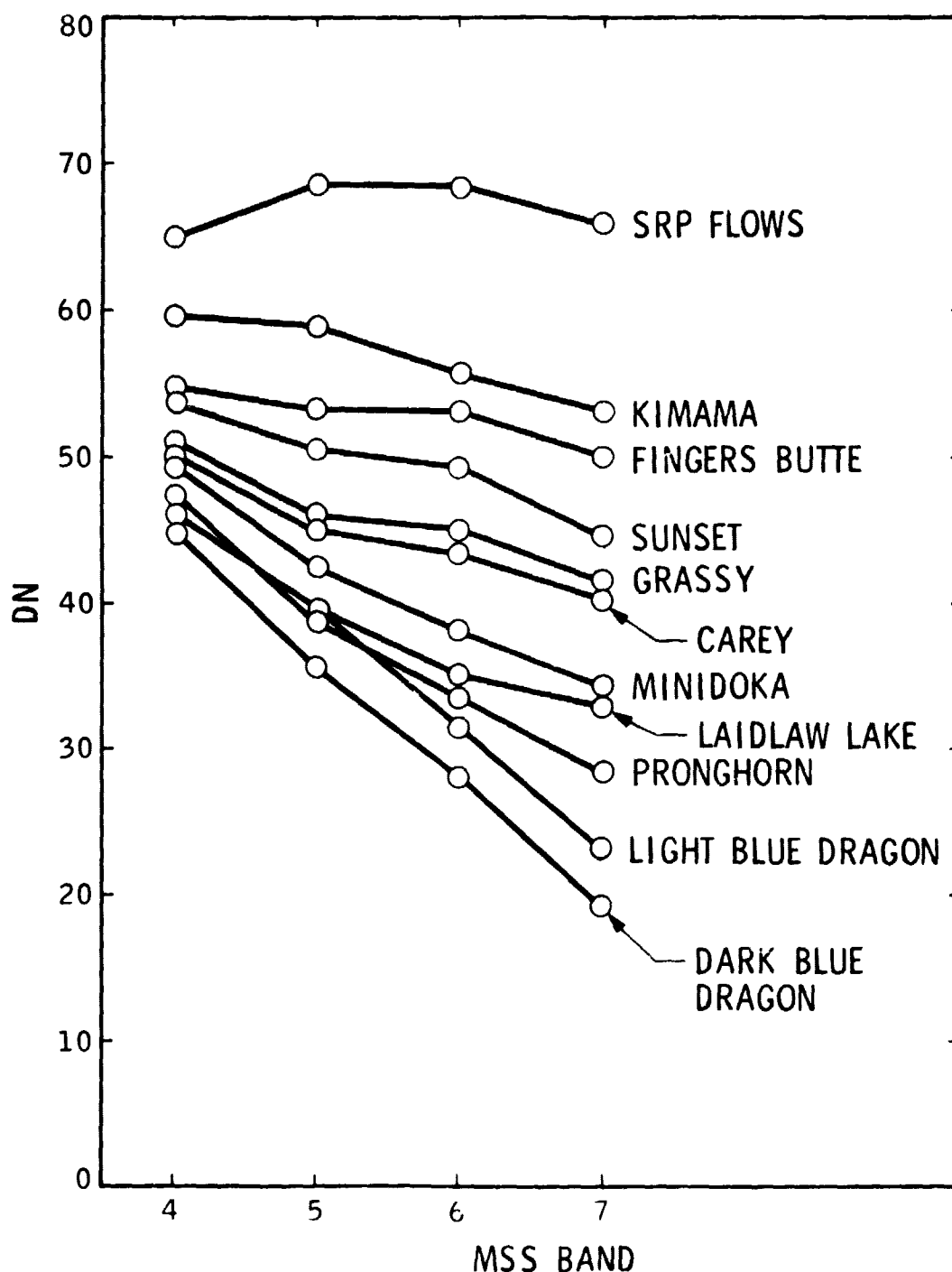


Figure 2. Digital number (DN) for each MSS band for the individual Holocene COM pahoehoe flows and the surrounding pleistocene SRP flows.

CRATERS OF THE MOON VOLCANIC FIELD, IDAHO

4 CLASS CLASSIFICATION

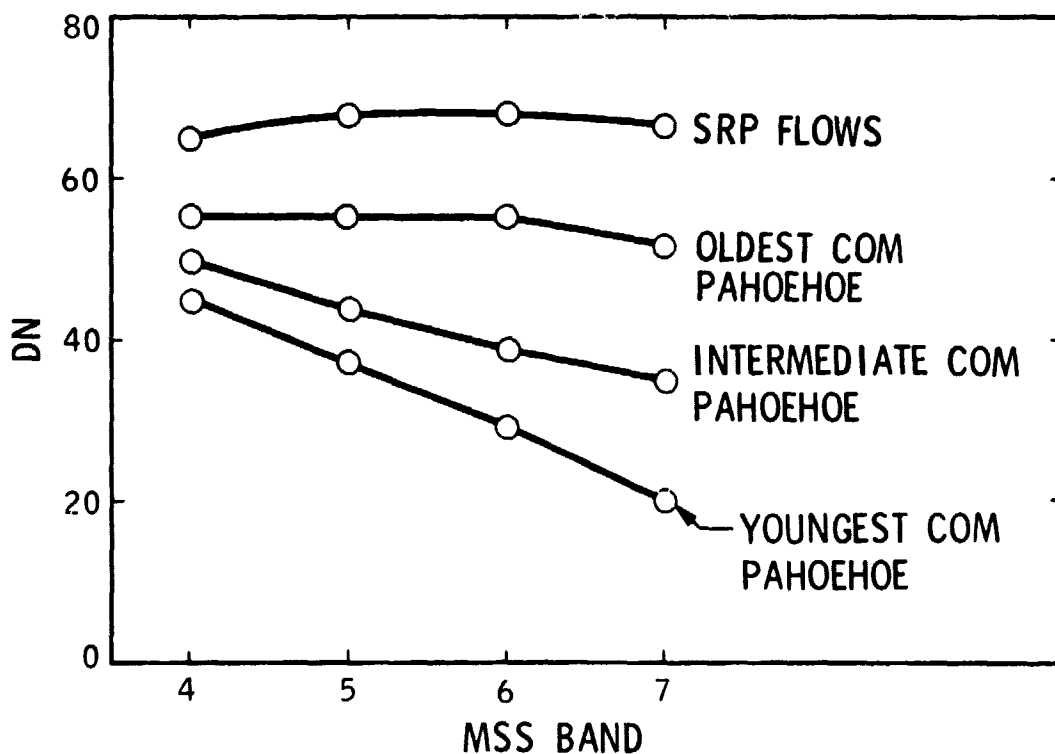


Fig. 1. Digital number (DN) for each MSS band for the three major age-related groups of Holocene COM pahoehoe flows and the surrounding Pleistocene Snake River Plain (SRP) flows.

PALEOMAGNETIC EVIDENCE FOR EPISODIC VOLCANISM ON THE SNAKE RIVER PLAIN

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Paleomagnetic measurements on basalt from Holocene and latest Pleistocene lava fields on the eastern Snake River Plain indicate that, in most cases, each of these lava fields formed in a short period of time. Typically, a single paleomagnetic field direction is recorded by all the rocks within a given lava field; this implies a total period of eruption shorter than several decades. The Craters of the Moon Lava Field, along the Great Rift, is an exception to this rule, however. Here many discrete episodes of eruptions have occurred from late Pleistocene to about 2100 years BP.

The direction of the geomagnetic field in the western United States is currently changing at rates between 4° and 5° /century. From the study of the remanent magnetization of hearths at archeological sites (Dubois, 1974) found that the direction of magnetization varied at an average rate of 7.5° /century between 1050 and 450 BP. The position of the Virtual Geomagnetic Pole (VGP) computed from the field directions at sites in southwest United States varied at an average rate of about 8° /century. A study by us of 12 ^{14}C dated lava flows in the western United States that range in age from 3000 to 1500 years BP suggests that the local field directions have varied at an average rate of 5.5° /century and the position of the corresponding VGP has varied at an average rate of about 6.5° /century. The observed range of inclination has been from 44° to 73° , and declination has varied from 343° to 24° . The direction of magnetization of a lava flow, in general, can be determined within 2° , at the 95 per cent confidence level, by the methods we are using. Differences in direction of magnetization corresponding to differences in age of 20 to 50 years generally can be resolved.

Six lava fields of Holocene and latest Pleistocene age, the Shoshone Ice Caves, Craters of the Moon, King's Bowl, Wapi, Cerro Grande, and Hell's Half Acre Fields are scattered over the eastern Snake River Plain. Lava flows in these fields are distinguished from older Pleistocene lava flows by the absence of superimposed loess or the presence of only a scattered thin veneer of loess and by the absence or scarcity of vegetation. From the air or from space these young lava fields are recognizable as striking dark patches on the Plain. Although the Shoshone Ice Caves, Wapi, Cerro Grande, and Hell's Half Acre Fields are each more than 100 sq. km. in area and include many flows, the directions of magnetization for individual flows within each field are not statistically discriminable. Eruption of lava at each field may have occurred in a few years or less. On the basis of ^{14}C dates of overridden plant material and correlation by direction of remanent magnetization, this short burst of eruptive activity took place at about 10,500 BP at the Cerro Grande Field, at about 4100 BP at the Hell's Half Acre Field and about 2400 BP at the Wapi and nearby King's Bowl Fields (Kuntz, personal comm.).

A much more complex and extensive sequence of lava flows is found at the Craters of the Moon Field. Different amounts of eolian silt and vegetative cover occur on different lava flows, and a stratigraphic succession of lavas is present which is fairly readily recognized and which can be mapped.

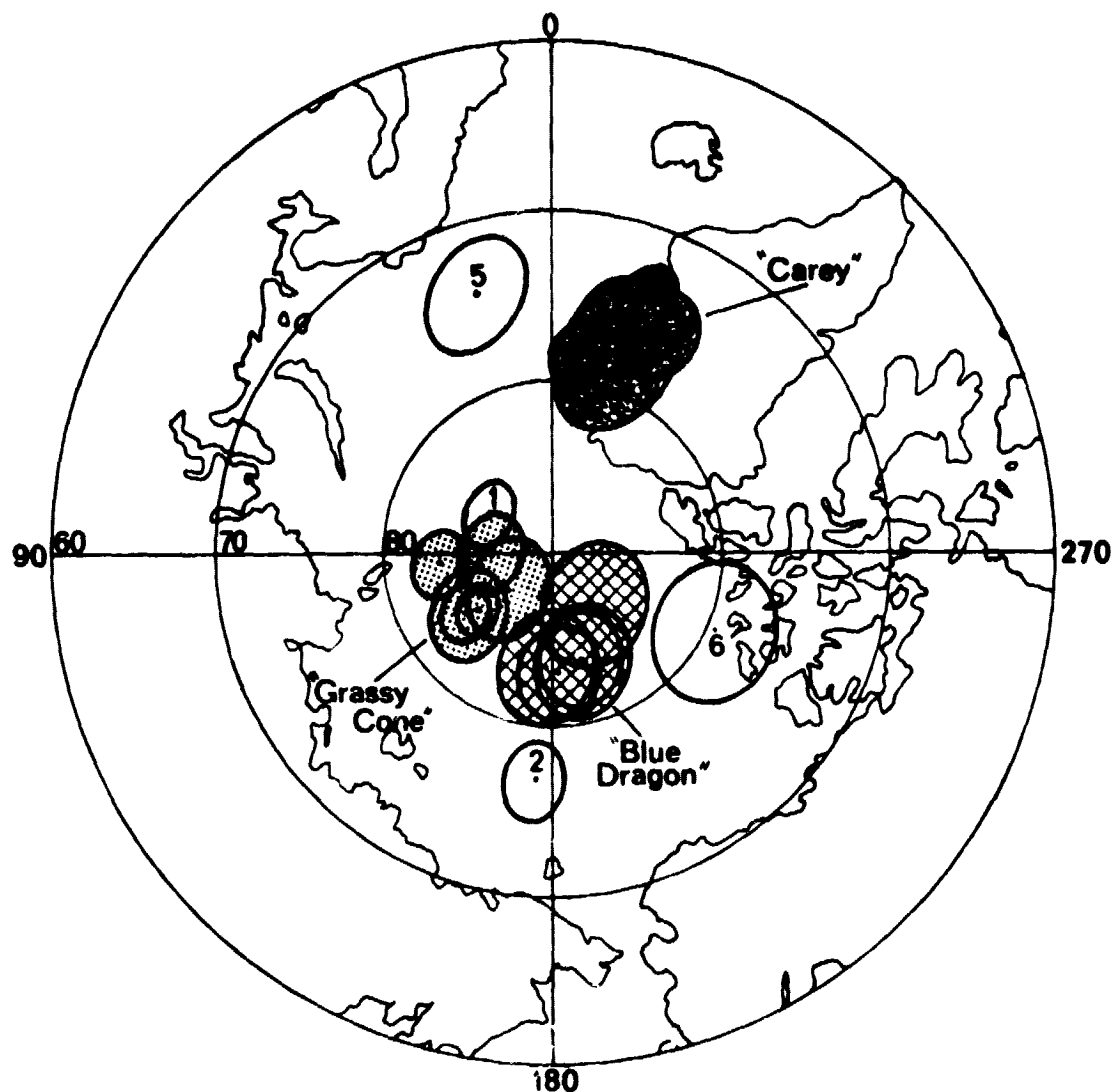


Fig.1 North polar plot showing virtual geomagnetic poles and ovals of 95 per cent confidence limits from lava flows in Craters of the Moon Lava Field, showing groups which document episodic behavior.

Available ^{14}C dates show that the age of this sequence spans more than 9000 years. Paleomagnetic measurements and stratigraphic studies carried out to date indicate that there have been at least 6 discriminably short episodes starting at about 11,000 BP. Figure 3, is a north polar plot of the VGP's and ovals of 95 per cent confidence, from these lava flows in the Craters of the Moon. Pole 2 is from an older pahoehoe unit which seems to be one of the oldest stratigraphically. In the second of these episodes, here designated the Carey episode, extensive pahoehoe lavas of the Carey Tongue were erupted. The direction of magnetization of the Northeast Sunset Flow, dated at 11,100 BP, falls on the direction of the Carey Tongue; this flow is believed to have erupted during the Carey episode. In the next episode a group of pahoehoe lavas were erupted, including a long flow from Grassy Cone which lies on top of the Carey Tongue lavas. Directions

of magnetization are tightly grouped for lava erupted during this episode, here designated the Grassy Cone episode. Two episodes followed in which aa lavas were erupted. Directions of the geomagnetic field were widely separated in these two episodes, shown as poles 5 and 6 on Fig. 1. They were also widely separated from the field directions at the time of the earlier eruptions of pahoehoe lava. Last in stratigraphic superposition are a group of very fresh appearing pahoehoe lavas, here designated the Blue Dragon group, for which four ^{14}C dates around 2100 BP have been obtained.

The Craters of the Moon lava Field sits directly astride the Great Rift, an extensional fissure system spanning the Snake River Plain from north to south. The Great Rift is parallel to the northwest trending fissure systems which seem to provide the structural control for the location of many of the cones and flows on the Plain. The extended period of eruption at Craters of the Moon may be due to its location on the Great Rift, but the Wapi and King's Bowl Fields are also on the Rift and they erupted over short periods. Location on a well developed rift is not therefore the only condition for extended volcanic activity, but it may well be a necessary condition.

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TOPOGRAPHY OF SMALL BASALT SHIELDS

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Terrestrial analogs are helpful for interpreting shield landforms on the lunar basalt plains. According to high-resolution Apollo photographs, the summit depressions of low domes on the Moon's mare surfaces (e.g., Cauchy Omega) are rimless, simple craters and not complex calderas. The long-prevalent interpretation of lunar domes as shield volcanoes resembling the large piles formed on the Earth's sea floor recently has given way to comparisons of the mare domes with smaller types of shields that have only simple summit craters and occur on subaerial basalt plains (Pike, 1974; Whitford-Stark, 1975; Head, 1976; Head and Gifford, 1977; Pike, in press). Two common varieties of small shield volcano, the "Icelandic" type and the "basalt-plains" (also "scutulum" or "Faroe Island") type, are compared here quantitatively with cratered lunar domes. The two specific objectives of this experiment are identification of reasonable Earth analogs of the lunar domes -- in both size and shape -- and recognition of any further subtypes of Icelandic and basalt-plains shields. In both cases the basis of comparison is a set of five measurements: edifice height, flank width, and summit crater diameter, depth, and rim-crest circularity (Fig. 1, Table 1). The rationale and method of the analog approach -- testing for geometric similitude by tabular or graphic analysis and multivariate statistics -- have been set forth in detail elsewhere (Pike, 1974, and in press).

The best-known distinction between classic Icelandic volcanoes (e.g., Skjaldbreidur) and basalt-plains volcanoes (e.g., Grandview Crater, Idaho) is slope angle of the flank (Table 1, Fig. 2). Icelandic shields typically attain slopes of about 7° , whereas basalt-plains shields usually slope at about 2° (Macdonald, 1972). The basalt-plains volcanoes also are much smaller in overall size. Certain very small shields (e.g., Mauna Ulu, Hawaii) that are not of the larger, Icelandic, type also have steeper flanks (nearly 6° , average). These compose an additional category, "small, steep shields;" the remaining basalt-plains volcanoes henceforth are termed "small, low shields" (Table 1).

Data from 87 cratered shields were available for this experiment: 17 Icelandic type, 10 small and steep, 54 small and low, and six lunar domes. Of the 81 terrestrial shields, 37 are on the Snake River Plain, 25 are in Iceland, 8 are in Oregon and Northern California, and 8 are Hawaiian. The lunar features are Herodotus Omega, "D-caldera," Cauchy Omega, the dome near Rima Aristarchus VIII, and the two small domes near the crater Maraldi B; data for the latter five are from Apollo photogrammetry. Measurements for all but a few terrestrial shields were made on recent topographic maps, mainly at 1:24,000 and 1:50,000.

Figures 3 and 4 and data in Table 1 suggest that lunar mare domes -- as a class -- resemble none of the candidate terrestrial analogs -- as a class. Although average base diameter of domes is close to that for Icelandic-type shields, their average edifice volume, crater size, proportion of crater volume to edifice volume, and slope of the flank are wholly disparate. Multivariate analysis of average data in Table 1 (Fig. 4), which eliminates absolute size as a basis of comparison, also shows that domes

Table 1. Mean Dimensions, Ratios, and Volumes for Cratered Shield Volcanoes ¹

Type of Edifice	n	Crater Diameter (m)	Crater Depth (m)	Edifice Height (m)	Flank Width (m)	Circularity	Height/Depth	Width/Depth	Diameter/Diameter	Height/Width	Depth/Width	Edifice Volume (km ³)	Edifice Vol./Crater Vol.
Lunar dome	6	1935	102	142	4076	0.41	1.7	2.1	0.073	0.053	0.035	3.58	20
Icelandic shield	15	475	33	480	1050	0.48	14.7	8.5	1.02	0.070	0.12	9.24	2400
Small shield, steep	10	190	23	70	700	0.43	3.0	3.7	0.37	0.120	0.10	0.05	100
Small shield, low	54	370	24	75	2200	0.40	3.3	6.0	0.21	0.64	0.034	0.04	250
Regular crater	25	421	31	66	2060	0.50	2.2	4.9	0.16	0.073	0.032	0.35	120
Irregular crater	25	316	19	87	2500	0.28	4.6	7.9	0.28	0.059	0.035	0.64	650

¹ All quantities except circularities and volume ratios are geometric means (to correct for skewness).

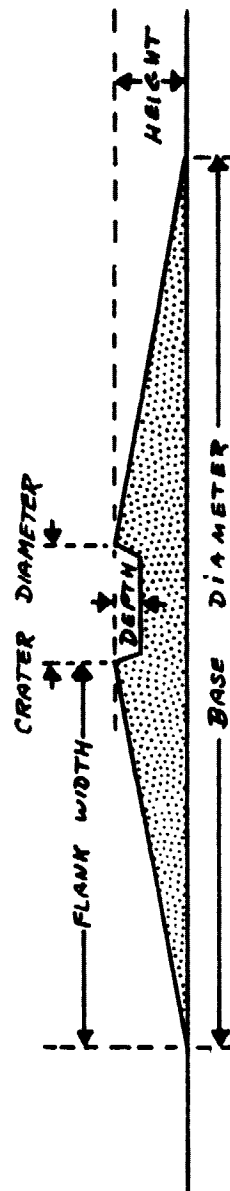


Fig. 1 Five topographic dimensions of a cratered shield volcano, defined diagrammatically. A sixth (dimensionless) property, circularity of the crater, is defined as the ratio of area of an inscribed circle to area of a circumscribed circle fitted to the outline of the rim crest.

are not like Icelandic shields or the smaller steep shields. Figure 4 does suggest that mare domes marginally resemble certain low basalt-plains shields in shape, as suggested by Head (1976), but differences still outweigh similarities.

Three main groups of volcanoes result from a cluster analysis (not shown) of all 87 craters and seven dimensionless properties (variables 5-11 in Table 1). The first group contains all but one Icelandic shield and all but one of the small, steep shields. Group two has 25 low basalt-plains shields; according to Table 1 and Figure 3, these volcanoes have a very small, shallow, and irregular summit crater. Group three contains all but one of the six lunar domes (Herodotus Omega is in group one, possibly the result of inadequate data) and 25 other low basalt-plains shields. Unlike the low shields in group two, the group three shields have much wider, deeper, and more circular craters (cf. Table 1, Fig. 3). Even these shields, however, still differ appreciably in shape from the lunar domes (Fig. 3, Table 1), which among other disparities have much wider craters relative to diameters of the volcanic piles.

Interpretation of these results in terms of the mode of origin of lunar mare domes could follow at least two lines of speculation: (1) lunar and terrestrial shield-forming eruptions basically are similar but some systematic perturbation has imparted a different shape to the lunar shields; (2) lunar and terrestrial shield-forming processes differ fundamentally and yield different landforms. As an example of the first approach, behavior of mare lava as a Bingham material (Moore and Schaber, 1975) may have drastically reduced the distance lava emitted from a dome can travel from the central crater and thereby reduced the width of the rim-flank vis-a-vis diameter of the crater. The convex shape of the Cauchy Omega margin supports this view. However, the diameters of summit craters on lunar domes far exceed diameters of craters on terrestrial shields (but are smaller than calderas on large shields), possibly a more fundamental difference. Perhaps this discrepancy resulted from extraordinary enlargement of the lunar summit craters by collapse during an earlier or more rapid withdrawal of the magma column beneath the vent than is typical of terrestrial shield eruptions. Alternatively, magma columns that fed domes on the lunar maria initially may have been broader than are those forming small shields on terrestrial basalt plains. Either of these possibilities suggests a fundamental difference in the development of the lunar domes. In any case, clear-cut terrestrial analogs of the lunar domes simply may not exist.

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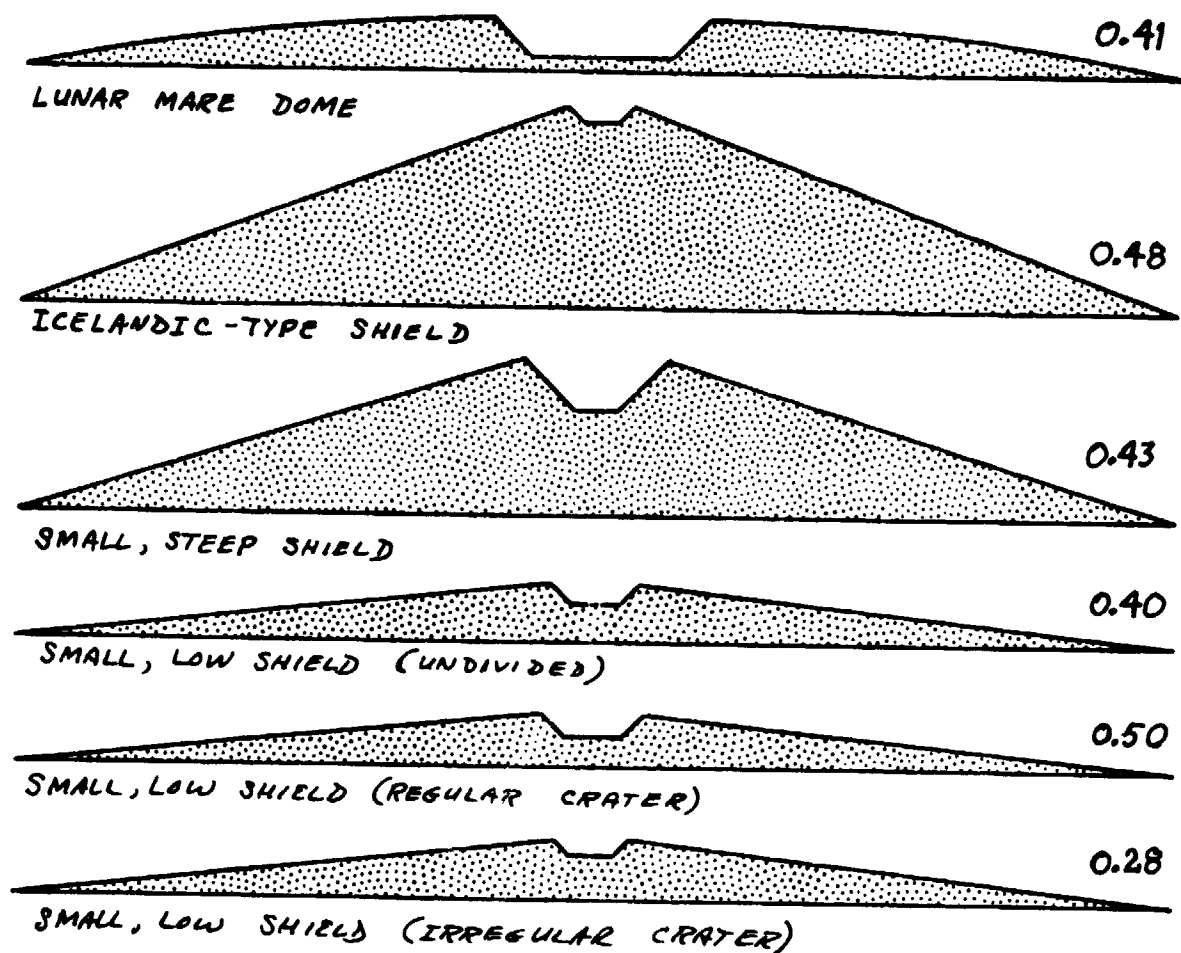


Fig. 3 Comparison of volcano shape (for size, see Table 1). Average topographic profiles for six types of basalt shields (from data in Table 1), scaled to the same base diameter. Curvature of flanks and crater interiors is arbitrary. Vertical exaggeration 3X. Numbers are mean values of rim-crest circularity for the summit crater. Except for slope and circularity, the lunar mare domes closely resemble none of the terrestrial basalt shields.

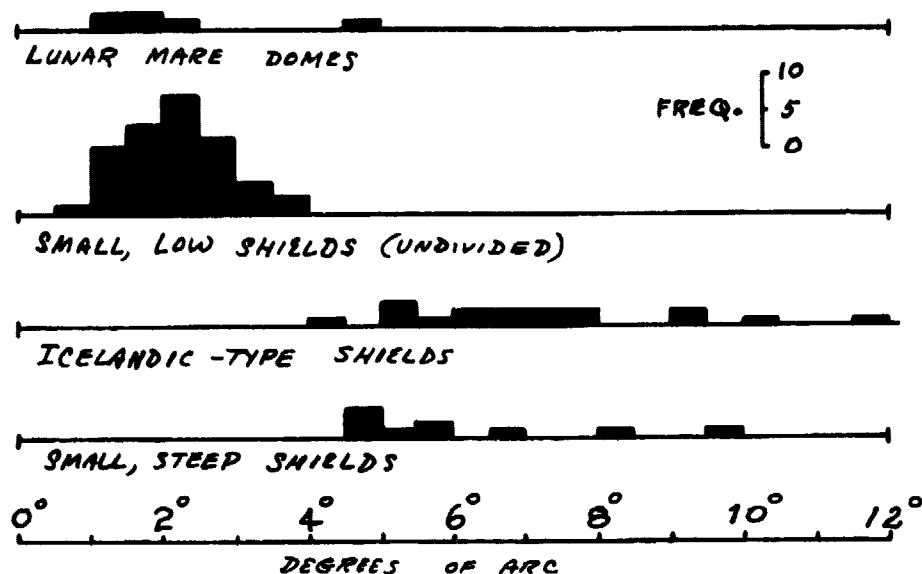


Fig. 2 Slope-frequency data for flanks of four types of cratered shield volcanoes. Small, low shields on basalt plains -- but not Icelandic-type shields or steep basalt-plains shields -- slope at about the same angle as flanks of lunar domes.

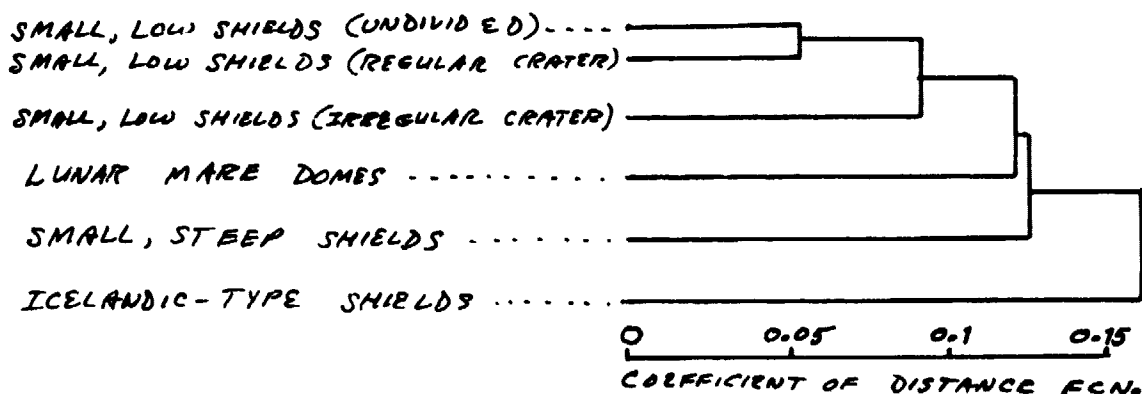


Fig. 4 A shape-classification of cratered shield volcanoes. Dendrogram displaying results of a cluster analysis for six types of basalt shields on three principal components (for explanation see Pike, 1974) that reflect seven original (averaged) ratios (variables 5-11, Table 1). Values of distance-function coefficient at which types of edifices cluster are low for similar types of shields and high for different types. Position of vertical lines along horizontal axis shows affinity at which each type of shield joined other types. There are no well-defined clusters of similar shields, but rather six different types of edifices. The lunar domes -- as a class -- closely resemble none of the terrestrial shields -- as a class.

INTERNAL STRUCTURE OF CINDER CONES

Crumpler, L.S. and J.C. Aubele, Department of Geology, University of New Mexico, Albuquerque, NM 87131

In the volcanic plateaus of the Mount Taylor and Cerros del Rio volcanic fields of New Mexico, cinder cones exposed on the surface, naturally sectioned on the peripheries and completely denuded in adjacent valleys show an excellent continuum of the internal structure of small volcanic vents and their relation to internal feeder intrusions.

Surface expressions of 2 m.y.± old cinder, cinder-and-spatter and spatter cones throughout New Mexico show a characteristic inward-dipping layering near their topographic summits, presumably erosion remnants of the original structure. The general structure of cinder cones consists of inward-dipping consolidated beds around the crater or vent zone and outward-dipping beds of variable materials on the flanks of the volcanoes (Fig. 1).

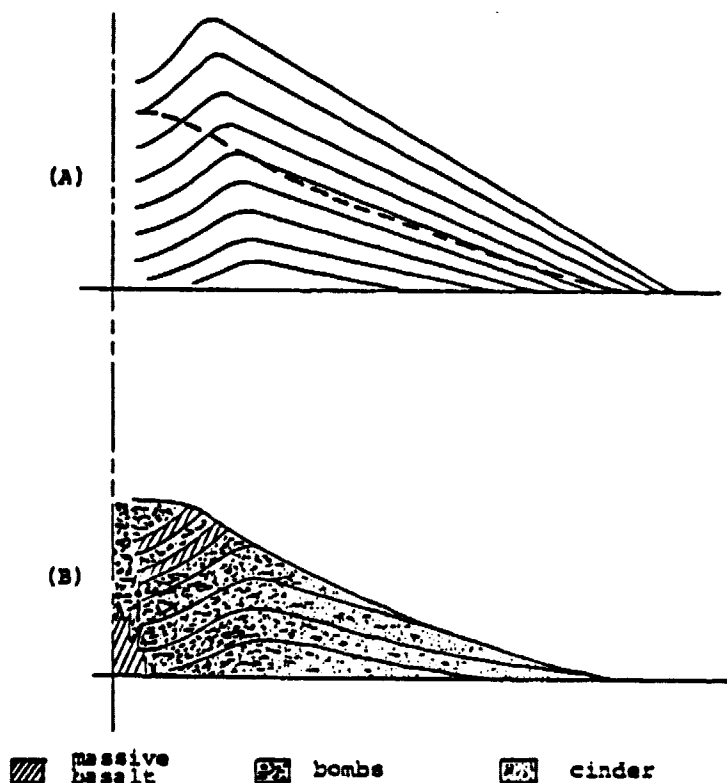


Figure 1. (A) Agglomerate and cinder layering in cinder-and-spatter cones showing the erosional profile (dashed line) of typical volcanoes. (B) Structure of a typical cone showing the massive basalt layers seen in the upper slopes, density of bombs and cinder and central dike-agglomerate feeder.

(1) Inward-dipping units: The majority of inward-dipping units examined to date are agglomeratic with distinct layering ranging in thickness from a few centimeters to several meters. These units dip inward toward the vent at angles of 30° to 45° , increasing in dip near the crater rim crests. Dominate materials are oxidized blocks and bombs cemented or welded in a matrix of ash, lapilli and fragments of basalt. They alternate in some cases with dense black units of basalt up to 2 m thick (Fig. 1b), which presumably are the remains of solidified lava pools. Some basaltic masses are possibly intrusive, but most are better interpreted as flows whose inward-dipping shapes are the results of confinement in the craters of the host cone (Fig. 2).



Figure 2. Vent 24 in the Mount Taylor field showing typical position of concentric and inward-dipping outcrops of massive basalt (arrow). Structural crater is breached to the south (left).

Where craters were breached during eruption, these basaltic masses show foliations closely paralleling the walls of the breach rather than simple concentric arrangements; these are probably the remnants of former lava levees. Summit craters which owe their origin to erosion acting along inward-dipping concentric units we have termed structural craters to distinguish them from primary craters, i.e., they are small erosional amphitheaters. (2) Outward-dipping units: Interbedded lavas, fine ash, cinders and sparse blocks typically form quaquaversal units dipping at angles up to 30° on the upper cone (where still preserved). Coarse-grained debris, including blocks, large cinders, bombs and basaltic boulders may represent slide and talus emplacement on the mid to distal slopes of cones. Bedding in this case is generally discontinuous and individual beds or channel slide deposits are well-sorted. Many cones show only quaquaversal layering and a central depression.

Coating of the outer slopes of the cinder cone with basalt and subsequent removal of soft inner cinders or explosive last activity and crater widening are thought to be responsible for this arrangement.

Where cones are naturally half-sectioned the relationship between feeder and vent is visible. Two types are recognized: (1) agglomerate-plugged vents and (2) basalt-plugged vents. In (1) agglomerate presumably filled the crater by insliding and backfall prior to the last invasion by basalt. The intrusion fingers out as dikelets at depth into the central agglomerate plug, in many cases forming a matrix for the coarser and more porous agglomerate. This is well illustrated in Cubero volcano, New Mexico. In (2) upwelling of basalt ceased while it still filled the vent forming a massive plug near the surface and filling the interior of the cone. This is best preserved in a cone almost perfectly sectioned in the west wall of Lobo Canyon near Grants, New Mexico.

Also peculiar to many dissected vents is a strong pattern of concentric or "onion skin" jointing in the deep or near-intrusion interiors which may be stoppe fractures. In some cases these have acted as channel for intrusion by basalt. In this way the main body of the intrusion works its way higher within the mass of the cone by stoping and engulfing portions of the cone agglomerate. Cinders and agglomerate in many of the deeply eroded volcanic necks of the Puerco valley often form collars around an interior of massive basalt and may represent stoped cone materials which were incorporated in the down-going convective collar of active intrusions.

REMOTE SENSING STUDIES OF ICELANDIC VOLCANIC FEATURES

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Studies of Icelandic volcanic features involving image processing techniques applied to Landsat multispectral scanner data and airborne radar are being pursued to better quantify the textural, compositional and temporal effects reflected in remotely sensed data (Lefebvre and Abrams, 1977). Advanced processing techniques, including the use of hybrid parallelepiped and Bayesian maximum likelihood algorithms, allow discrimination of vegetation and surface texture, while 25 cm wavelength (L-band) radar provides additional information on surface roughness. Field correlations will be reported.

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ROLE OF LAVA TUBES IN FLOWS FROM OBSERVATORY VENT, 1971 ERUPTION OF MOUNT ETNA

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The 1971 eruption on Mount Etna started on April 5 near the foot of the Summit Cone at about 2985 m elevation. During the first phase of the eruption three distinct vents were in operation in this region; the Observatory Vent was the major one and opened just upslope from the Volcano Observatory which was destroyed by the lava. During the later phases of the eruption the center of activity migrated eastwards (Rittmann *et al.*, 1972); the eruption ceased on June 12, 1971.

The total length of lava from the Observatory Vent was about 3.7 km; the lowest elevation reached by the lava was about 2,200 m. The lava was erupted from boccas at the southern foot of the Observatory Vent cone and also from small boccas downslope from this cone (Fig. 1).

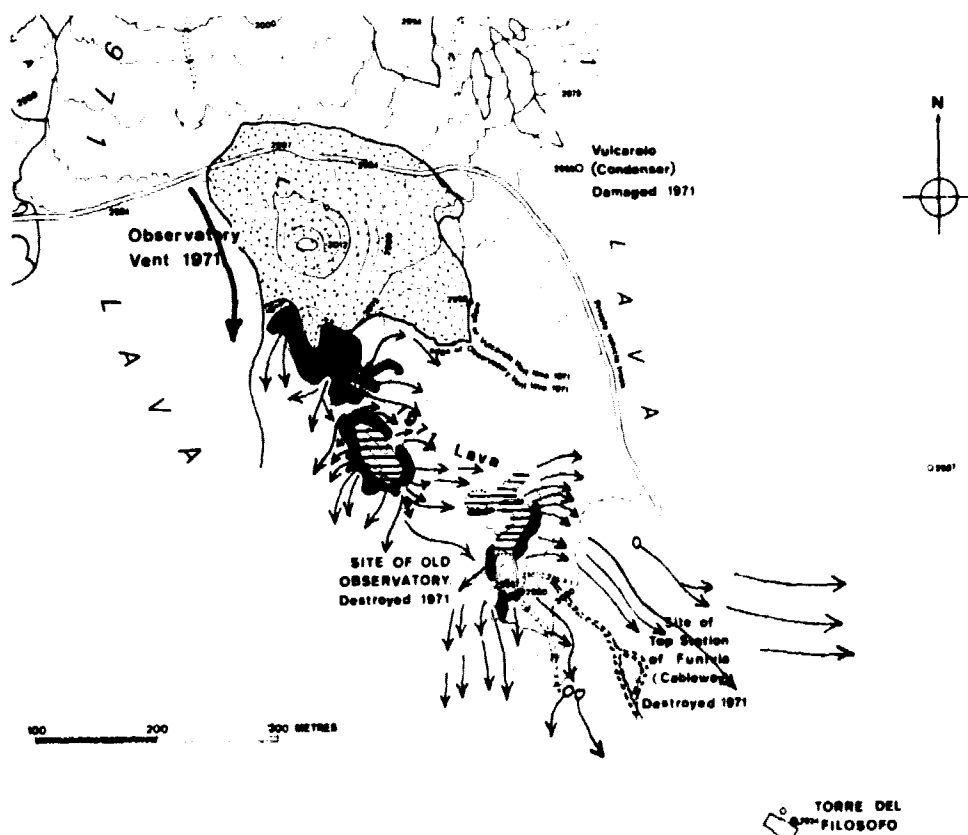


Fig. 1 Location map showing cone of Observatory Vent (stippled pattern), buff-colored altered lava (black) and collapsed tumuli (horizontal line pattern). Paths of small flows shown by arrows; larger lava channels shown by parallel dashed lines.

Immediately south of the cone is a well defined ridge of buff-colored blocks of altered lava (marked in black on Fig. 1). This ridge appears to mark the main fissure vent, but the distribution of the blocks suggest that the surface of the lava flowing from this area congealed and formed large tumuli which later collapsed (collapsed areas marked by horizontal lines on Fig. 1). The surrounding areas are covered by numerous flows of unaltered pahoehoe and aa lava that emptied from small boccas and cracks in the tumuli, and the collapsed material in the tumuli also is coated on its upper surface by unaltered lavas.

This pronounced ridge of tumuli continues for about 150 m from the foot of the cone; at the southern end it is covered by unaltered lavas erupted from cracks in the tumuli. About 60 m to the east the feature reappears as a more subdued channel floored with collapsed lava slabs, suggesting that it was originally a tube whose roof fell in at the end of the eruption. This channel turns sharply to the south where it is again bordered by uptilted blocks of altered lava and terminates in a pit which is about 40 m long and 30 m wide. This pit appears to have been an open pool of lava at least in the latter stages of the eruption. Overflows and squeeze-outs of lava border the whole of this channel and the pit.

From these observations it appears that much of the lava erupted near the Observatory Vent initially flowed in a large channel, part of which was the fissure vent itself. The channel became roofed over and the lava continued flowing in the tube causing alteration of the surrounding lava to give it a buff colour. Increase in the volume of lava being erupted caused uparching of the roof of the tube to form tumuli, and lava was emitted from cracks in the tumuli to give the present surface pattern of lavas as shown in Figure 1. Changes in level of lava in the pit caused repeated overflows to give lava channels (parallel dashed lines on Fig. 1) radiating from the edge of it.

The pit itself is draped with lava that remained on the surface when the pit finally drained. Alcoves in the wall of the pit suggest original openings to lava tubes initially fed from lava in the pit. The largest of these alcoves is near the southern end of the pit immediately below the surface of a channel. The channel itself is more than 100 m long and has a depth of about one meter; a few meters from the edge of the pit the channel bifurcates downwards and enters a lava tube (Croissant, 1972). This lava tube does not follow the line of the surface channel and has a total length of about 50 m (Fig. 2). From the point of entry of the overlying channel, the tube extends northwards as two small tubes which terminate near the wall of the pit (Fig. 2).

From these observations we conclude that lava flowed from the Observatory Vent tube/channel system into the pit where it escaped downslope through small lava tubes. Increases in the level of lava in the pit caused overflow into surface channels. At least in the case observed here, lava in the channel was directly connected with flowing lava in a tube below the surface. Downslope from this region there are a number of small boccas, and it is likely that the tube studied, as well as others not open to the surface, fed these ephemeral boccas.

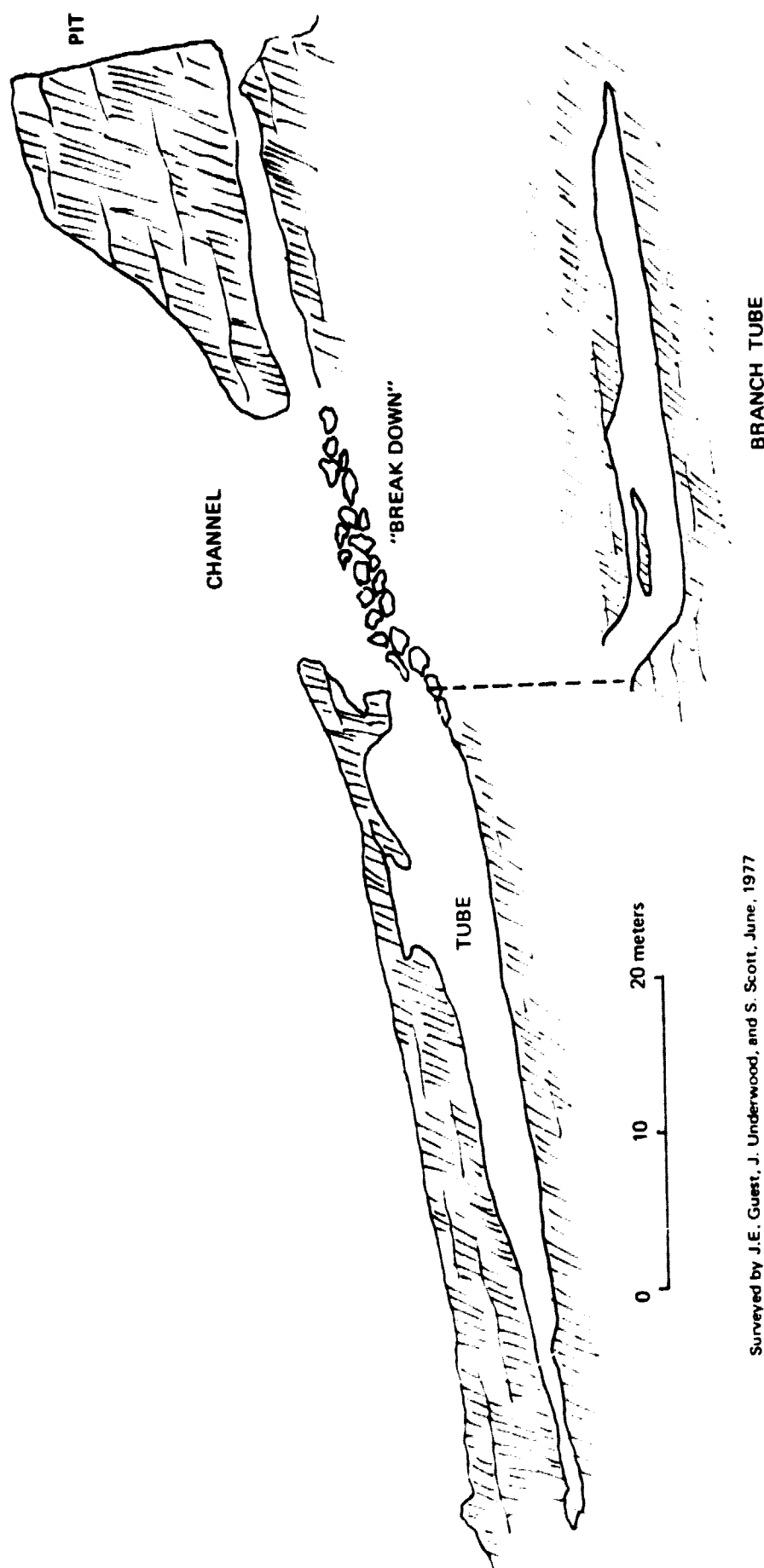


Fig. 2 Longitudinal cross section of lava tube in flow from Observatory Vent. North is to the right.

It is normal during eruptions on Mount Etna for aa lava flows to be fed from ephemeral boccas downslope from the main vent. The position of these boccas changes from day to day. On the basis of our study of the flows from the 1971 Observatory Vent, it is concluded that: (1) ephemeral boccas on Mount Etna are fed by lava tube systems emanating from the principal vent and (2) lava tubes may play a much more important role in the eruption of aa lavas than has previously been thought.

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POSSIBLE LUNAR ANALOGS TO SNAKE RIVER PLAIN BASALT MORPHOLOGY

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The lunar maria are generally devoid of well-defined lava flow fronts, the Imbrium flows representing notable exceptions. However, other features and textures on the maria resemble primary flow morphologies found in terrestrial lava fields including those in the Snake River Plain. Pahoehoe basalt flows typically exhibit a ropy texture at fine scales that could not survive the long-term degradational processes on the Moon. Nevertheless, such flows also exhibit broader scale features (dimensions larger than 50 m) that might survive meteoritic erosion on the younger lunar lavas. These features include:

1. tumuli and irregular knobs: maximum size approaching 50 m.
2. circular collapse depressions: ranging from 10 m to 50 m.
3. irregular plateaus: ranging from less than 50 m across to greater than 500 m.
4. irregular depressions: ranging from less than 50 m to greater than 500 m.
5. ridges: from very narrow widths to 20 m.
6. linear depressions (moats) between pre-existing relief and subsequent floors.

The association of these various forms produces a distinctive hummocky surface in several Snake River Plain basalt fields.

Similar hummocky mare surfaces occur on the Moon and are most visible under low solar illumination. Such hummocky mare units commonly contain numerous interlinking irregular depressions and exhibit irregular patches of smooth-surfaced units. Although a few regions display a directional trend, more characteristically a surface grain is absent. Moreover, major sinuous rilles crossing such mare units are notably absent or rare.

The mare-flooded interior of the Flamsteed Ring exhibits a similar hummocky surface that has been interpreted by Schaber et al. (1976) as highly degraded lava termini. Close inspection of this surface, however, reveals that the irregular scarps comprise the borders of irregular plateaus or depressions, which resemble the features in certain Snake River Plain basalt fields. In this analogy, the irregular scarps are not the termini of individual flows but are remnants of differential subsidence within a single flow unit.

Also within the Flamsteed Ring are numerous, small (50 m - 150 m diameter) rings characterized by a narrow depression encircling a low-relief mound (Schultz, 1976; Schultz and Greeley, 1975; Schultz et al., 1976). Such structures are called ring moats and are common within or adjacent to

hummocky mare regions and mare regions with low-relief irregular plateaus and depressions. Ring-moat structures fall into three classes: moat surrounding a dome; moat surrounding a dome with subdued summit depression; and moat surrounding an interior mare surface. To date, eight mare regions have been found that display ring-moat structures and generally correspond to young (Eratosthenian age) titanium-rich mare basalts. Isolated examples also are recognized in other regions. The great abundance of such features is illustrated by their spatial distribution within the mare-flooded crater Letronne in southern Oceanus Procellarum (Figure 1).

Several origins for these features have been postulated (Greeley and Schultz, 1975; Schultz and Greeley, 1976) and fall into two classes: origins at the time of mare emplacement (primary surface features) and origins at later times (secondary features). The preservation of irregular depressions and plateaus of comparable scale suggests that such structures were formed contemporary with the emplacement of these units. As primary flow features, ring moats may be produced by the terminus of thin (10 m) basalt flows surrounding pre-existing relief such as volcanic cones and tumuli, a process that can be documented in the Snake River Plain basalt fields (Greeley and King, 1975). On the Moon, the central relief of larger ring moat structures (diameter greater than 300 m) commonly exhibit central depressions, which may represent volcanic cones. The relief within smaller ring moat structures, however, generally lack such central depressions and may correspond to tumuli. Other possible mechanisms of ring-moat formation include autointrusions within thicker (50 m) flow units (e.g., see McKee and Stradling, 1970) and the inundation of impact craters. The latter process is believed to be responsible for ring-moat structures approaching 0.5 km to 1.0 km in diameter.

Both ring-moat structures and the textured mare surfaces may be providing important clues to the style of mare basalt emplacement where distinct flow termini are absent. By analogy with the Snake River Plain basalts, certain lunar flows represent plains-basalt eruptions characterized by relatively thin (10 m - 20 m) multiple flows, low flow viscosity, and numerous local vents (Greeley, 1975, 1976). Moreover, the close association of such regions with titanium-rich mare units recognized in the multi-spectral imagery of McCord et al. (1977) suggests that this was a common late eruptive style of the western lunar maria.

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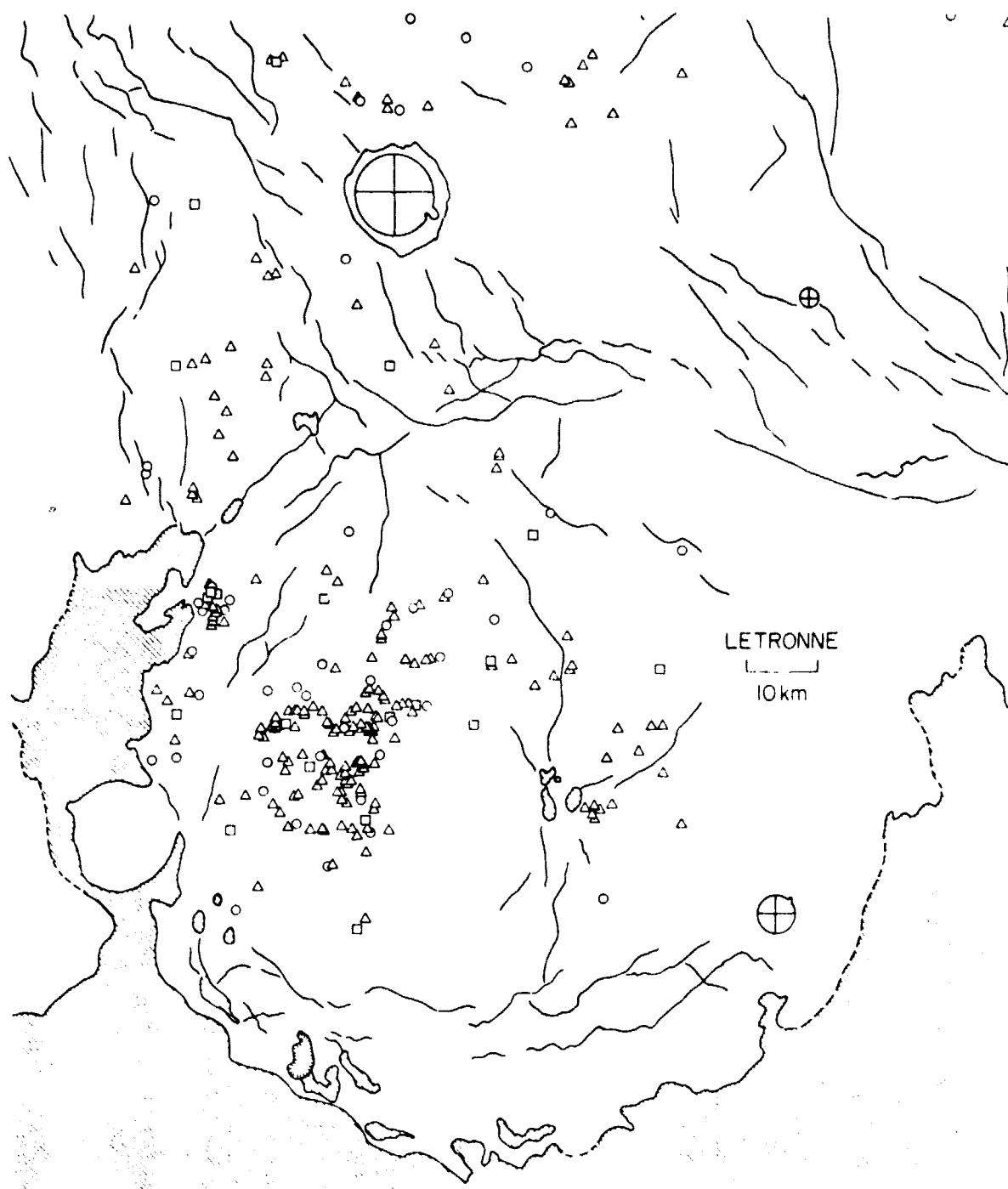


Figure 1. The distribution of ring-moat structures within the mare-flooded crater Letronne. Diameter of rings range from 30 m to 500 m; widths of moat range from 20 m to 100 m. Circles indicate ring moats; triangles, domes surrounded by moats; and squares, pitted domes (cones) surrounded by moats. Circles with interior cross locate craters larger than 1 km in diameter. Sinuous lines correspond to wrinkle ridges. Hatchured regions indicate non-mare (crater rim, highlands) surfaces.

LAVA FLOW MORPHOLOGY, DISTRIBUTION AND ERUPTIVE HISTORY WITHIN THE THARSIS REGION OF MARS

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Viking Orbiter image data of the Tharsis Montes region of Mars are currently being used to describe the morphology, distribution and eruptive history of lava flow materials from the giant volcanic shields. Five detailed lava flow scarp maps covering a total Mars surface area of $5 \times 10^6 \text{ km}^2$ have been prepared thus far on Viking orthophoto maps including the new 1:1,250,000 scale orthophoto subquadrangle maps being prepared by the U. S. Geological Survey (Flagstaff, Arizona). Picture element (PIXEL) resolution of Viking images used thus far range from 75m to 250m.

The volcanic materials from individual eruptive sequences are being delineated by flow morphology (including flow thicknesses), surface albedo and superposed crater density. Lava sources are difficult to recognize for all but the youngest flow sequences but numerous eruptive vents have been recognized. Maximum flow lengths exceed 1000 km.

One of the most prolific source region for mappable lava flows appears to be the summit and flank regions of Arsia Mons. Flank eruption sources for this shield extend out from the summit a distance of over 1600 km. Summit eruptions for Arsia Mons are characterized by thinner (10-20m) narrow, channeled flows due to steeper gradients (0.008) and reduced eruption rates (Carr et al., 1977) while the flank eruptions are characterized by thicker (20-60m), wider, non-channeled flows deposited on lower gradients (0.004-0.003) and associated with considerably higher eruption rates. The youngest volcanic sequence, emanating from a 160 km long fissure of the south side of the Arsia Mons summit caldera, covers a total area of $5.6 \times 10^5 \text{ km}^2$ and represents a minimum volume of $3 \times 10^4 \text{ km}^3$. This volume is of the same order of magnitude as the total amount of lava extruded onto Mare Imbrium during the last 2+ billion years (Schaber, 1973). The earlier eruptive sequences associated with the Arsia Mons shield appear to have been of similar or greater magnitude.

A minimum of seven massive eruptive sequences associated with the west to south-east quadrants of the Arsia Mons shield have been identified thus far. Flow thicknesses ranging from 20m to 60m on the lower Arsia Mons slopes (0.004-.003) were found to represent yield strengths from $6.0 \times 10^2 \text{ Newtons/m}^2$ to $1.8 \times 10^3 \text{ Newtons/m}^2$ and possible SiO_2 contents of from 43 to 48 percent; using the techniques described by Hulme (1974), Moore and Schaber (1975) and Hulme and Fielder (1977). Forty Viking Orbiter frames of Arsia and other shields of the Tharsis Montes have thus far been counted for craters larger than 1 km diameter. These data are being utilized with photogeologic superposition relationships to verify stratigraphic sequences of eruptive materials.

The technique developed by D.W.G. Arthur for measuring lava flow heights from the Viking Orbiter images depends on readouts of the original spacecraft records and combining of pixels in either columns or rows depending on the direction of the shadow. The three complications are noise, picture blurring and the extremely small dimensions (one pixel or less) of some of the shadow widths. Approximate techniques have been developed for the first two. The last implies certain

rather natural limitations for very low objects such as lava flows. However, lava flow measurements are thought to be good to within 10 to 20 percent using this technique.

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LAVA EROSIONAL CHANNELS ON MARS

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Bahram, Vedra, Maumee and Maja Valles are a series of east-west trending valleys on Mars, located 500 km west of the landing site of the Viking I (VL-I) spacecraft. They were first recognized in images obtained by the Viking Orbiter 1 (VO-1) spacecraft which revealed sinuous 0.5 to 5 km wide depressions, termed rilles in this paper, connecting the plains of Lunae Planum and Chryse Planitia. Portions of the Bahram Vallis bear a morphological resemblance to sections of certain of the larger lunar sinuous rilles such as Schroters Valley and Hadley Rille. The rilles of Maumee, Vedra and Maja Valles, however, have a complex anastomosing character for which there is no lunar analog. Maja Vallis has also been modified and eroded by large scale flow phenomena, as described by Cutts and Blasius (1977a).

Evidence for the formation of Bahram Vallis by erosive flow along the rille axis consists of : 1) concordance with Chryse Planitia, which suggests plains materials were transported along the rille into the basin, 2) variations in morphology along the rille which correlate with topography in a manner which suggests incision by flow of a dense fluid; the rille is broad and shallow meandering on the flat plateau, but it becomes narrower, deeper, and nearly straight passing through mountainous terrain. 3) Internal structure in the flat-floored section, possibly an inner nested channel. In Vedra and Maumee Valles this evidence is supplemented by 4) the existence of a complex of anastomosing valleys which emerge from the upper plain and converge to form wider valleys which are concordant with the lower plain. 5) The occurrence of 'hanging valleys' which are distinctive fluid erosional features resulting from vally capture. No bedforms which definitively prove fluid erosion have been identified, and segments of these various valleys could have originated tectonically. However, the weight of the evidence favors a fluid erosional origin.

Among possible mechanisms of fluid erosion, wind is totally inconsistent with rille morphology; ice and water are compatible with some aspects of rille morphology but are rejected because of the lack of morainal or delta deposits. Lava erosion is also compatible with rille morphology and with the deposits observed at the mouths of the rilles which are widely accepted to be lava. Lunar analogy also favors a volcanic explanation over other mechanisms, but definitive evidence of a lava erosional origin is lacking at present because of the absence of high resolution imagery of the channel floor and because of the lack of a generally accepted theory of the relationship between the sinuous ridges that characterize the plains of Chryse Planitia and the mode of basalt emplacement.

Crater age data further constrain the likelihood of alternative mechanisms of rille formation. Crater density measurements and crater superposition/

intersection relationships place the age differential between rilles and surrounding volcanic plains at 0.22 and 0.24 respectively at the 95% confidence level. Penecontemporaneity in age supports a volcanic origin for the rilles through two different lines of argument. Firstly, it is a necessary condition for both plains and rilles to have been formed by lava. Secondly, it requires a remarkable coincidence for an erosional episode involving some other high density flowing medium to have occurred just once in Mars history and almost contemporaneously with the effusion of plains basalts.

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THE SMALLER CENTRAL VOLCANIC CONSTRUCTS OF THE THARSIS REGION OF MARS

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In addition to four great shields and Alba Patera at least seven smaller central volcanoes occur in the Tharsis region of Mars (Carr, 1973). Viking images show these features to be embayed for the most part by adjacent plains. They thus seem to represent an earlier period of volcanism. These features are much more highly varied in character than are the shields. Flank slopes vary from strongly convex upward to nearly uniform and summit caldera represent from 5% to 90% of the areas of individual constructs. The character of flank surfaces is highly variable also. Some appearing relatively smooth while others are densely cratered. Fault traces on the volcanoes are commonly cut off at the contact with the younger plains flows, so the volcanoes may serve as windows through which to view earlier patterns of faults now for the most part buried.

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VOLCANISM IN THE CRATERED UPLANDS OF MARS A PRELIMINARY VIKING VIEW

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Viking Orbiter systematic mapping has given new insight into the geologic evolution of the southern hemisphere of Mars. Although coverage is incomplete and of non-uniform quality, preliminary mapping has been completed for volcanic terrains on Mars. Mariner 9 results showed that most of the southern hemisphere of Mars is heavily cratered (termed the *cratered terrain hemisphere*, which makes up 55 percent of the martian surface), whereas much of the northern hemisphere has been extensively modified and resurfaced (Mutch and others, 1976, and references therein). From Mariner 9 and data from previous missions, it was observed that most large martian craters have been extensively modified and the heavily cratered terrain seen on the moon has no martian equivalent (Murray and others, 1971; Wilhelms, 1974). Wilhelms (1974) suggested that extensive volcanism has modified much of the ancient cratered terrain of Mars, producing smooth intercrater plains. Viking photographic coverage has produced supporting evidence for this conclusion, and provides details on the styles of volcanism which modify the cratered terrain. Preliminary mapping based on Viking data for volcanic terrains on Mars shows the following units as important modifiers of the martian cratered terrain hemisphere:

1) *Patera*. The patera, of which Tyrrhena, Hadriaca, and Amphitrites are most prominent, occupy approximately 0.3 percent ($237,000 \text{ km}^2$) of the cratered terrain hemisphere. These landforms appear to be shield volcanoes in different states of degradation, ranging from the relatively young Tyrrhena Patera to the nearly obliterated patera of the Amphitrites region. A small, deeply dissected patera-like construct ($-200, 1870$) has recently been found (Greeley, 1977) and many more may exist in regions for which poor photographic coverage exists. These constructs all lie stratigraphically above the martian intercrater plains (Scott and Carr, in press).

2) *"Plains" style volcanism*. "Plains" volcanics are characterized by the presence of lava tubes, lava channels, cones, domes, low shields, and leveed flows (Greeley, 1976). These volcanics comprise approximately 2.9 percent ($2.3 \times 10^6 \text{ km}^2$) of the cratered terrain hemisphere, being most prominent around the margins of the Hellas basin. Numerous cones and low shields occur in the vicinity of Amphitrites Patera and must have developed much later than the highly degraded patera. Detailed mapping of a "plains" area east of Hellas suggests at least some of these cones and low shields may be contemporaneous with smooth plains layered deposits, suggesting the layered plains are composed of a succession of individual lava flows. The close association of many of these "plains" volcanics with the larger patera attest to a prolonged volcanic evolution.

3) *Flood-type volcanism*. Large areas of plains are characterized mostly by flow scarps and wrinkle ridges, typical of flood-type volcanism. These units are most extensive as basin-fill materials and comprise about 4.7 percent ($3.7 \times 10^6 \text{ km}^2$) of the cratered terrain hemisphere. Although flow fronts are abundant in this unit, aeolian cover frequently masks primary volcanic surface morphologies, particularly in active aeolian regimes such as the floor of the Hellas basin. Because of this mantling, the full areal extent of the flood lavas is uncertain.

4) *Plateau plains*. The plateau plains (unit *plc* of Scott and Carr, in press) comprise almost 36 percent ($28.5 \times 10^6 \text{ km}^2$) of the cratered terrain hemisphere and about 20 percent ($28.8 \times 10^6 \text{ km}^2$) of the entire surface of Mars. It was observed from Mariner 9 data that the densely cratered terrain of the lunar highlands has no equivalent on Mars and it was suggested the smooth intercrater plains that occur over much of the cratered terrain hemisphere were volcanic in origin (Wilhelms, 1974). Viking photography supports this conclusion. Wrinkle ridges

are characteristic of this unit. Flow scarps of probable volcanic origin are also widely distributed in the plateau plains. Many of the plateau plains are intimately associated with volcanic centers of the southern hemisphere and appear to comprise the basal stratigraphic unit of the Amphitrites Patera volcanic complex. This unit is distinguished morphologically from the cratered terrain (unit *hc* of Scott and Carr, in press) which frequently displays fine-scale channeling seldom seen on the plateau plains unit. These plains may be volcanic in origin, although unequivocal planetwide evidence for this origin cannot be found. The widespread occurrence and stratigraphic relations of the plateau plains suggest extensive volcanism, possibly contemporaneous with the end of the period of early, heavy bombardment (Soderblom and others, 1974). Crater studies (summarized in Chapman and Jones, 1977) strongly argue for an episode of extensive crater obliteration and, although other photogeologic evidence (e.g., small scale channeling in cratered terrains) suggests earlier intensive erosion on Mars, the eruption of the plateau plains volcanics may have been important in the removal of small (less than 10 km) craters in the cratered terrain hemisphere.

In summary, Viking mapping shows that the cratered terrain hemisphere of Mars has been extensively modified by volcanic processes. The earliest discernible episodes of volcanic modification involved emplacement of the plateau plains unit, possibly coincident with episodes of planetwide crater obliteration (see Chapman and Jones, 1977, and references therein). This early volcanism was pervasive (Wilhelms, 1974) and extensively modified the ancient primordial crust. Flood volcanism, possibly accompanied by shield-building episodes (e.g., patera) followed; flood-type flows are roughly correlative with the "cratered plains" units of Lunae Planum. "Plains" style volcanism is the youngest *recognized* episode of volcanism in the cratered terrain hemisphere; "plains" volcanic centers are isolated and confined to areas of previous volcanic activity. All volcanic units have been subjected to subsequent aeolian erosion.

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NON-BASALTIC SHIELD VOLCANOES

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The volcanoes Fuji and Kilauea have similar basaltic compositions, but completely different morphologies. Other volcanoes, with similar morphologies, may have widely different chemical compositions. These observations illustrate that although morphology and morphometry are generally considered to be good guides to the petrology and eruption styles of volcanoes, other rock types and eruption mechanisms can also produce remarkably similar landforms. As an example of the variety of ways a single landform can be constructed, and as a guide to the interpretation of shield volcano morphologies, this paper (1) summarizes the characteristics of basaltic shields, (2) documents the morphologies and eruption mechanisms of other shields composed of more felsic rocks, (3) compares basaltic and felsic shields, and (4) re-examines interpretations of shields on the Moon, Mars, and Venus.

Basaltic shield volcanoes. Circular piles of basaltic lavas with slopes of $1/2^\circ$ to 10° are common in many volcanic provinces. These basaltic shields are built by frequent eruptions of moderate to small volumes of fluid lavas from central vents and/or flank fissures. The smallest shields have basal diameters (D_s) of a few kilometers, whereas for Hawaii and other large shields, D_s is commonly a few hundred kilometers. Shield heights (H_s) are roughly $1/10 D_s$, and collapse calderas occur at their summits. Calderas are often nested or intersect, and Nordlie (1973) documented trends in caldera circularity and overall diameter (D_c) for the Galapagos shields. For large shields there is no consistent relation between D_c and D_s , but for shields smaller than about 15 km in diameter, D_c is approximately equal to 5% of D_s . The flanks of basaltic shields are surfaced by well-defined lava flows that often show lava channels and tubes.

Felsic shield volcanoes

Lava shields. Fluid flows of lavas commonly considered more silicic than basalts occasionally form cones with slopes as low as those characteristic of basaltic shields. Two examples are the phonolite shields, Suswa and Kilombe, within the Kenya Rift Valley. (Phonolites contain ~55% SiO_2 , ~13% alkalis, and ~1% MgO). Early eruptions of Suswa built an elongated shield with D_s 20x17 km, H_s ~530 m, and average slope (S_s) $\sim 3^\circ$ (Johnson, 1969). The first shield-building flows were thick (6 to 25 km) and voluminous, but later flows were not as extensive. A caldera (D_c 12x8 km) formed by collapse, and afterward voluminous pumice eruptions from ring faults mantled the shield.

Kilombe is a smaller and steeper rift volcano (McCall, 1964). Elongated parallel to the rift valley, the long axis of Kilombe is ~10 km, $H_s \approx 450$ m, and S_s averages $\sim 7\ 1/2^\circ$. An irregular caldera ~3 km wide is floored by pumice tuff, but there are no other pyroclastic deposits outside the caldera, and McCall believes that the crater formed as a result of deep-seated cauldron subsidence. The phonolite lavas that built Kilombe erupted from rift-aligned fissures, not from a central vent as in the case of Suswa.

Lava and pyroclastic shields. A series of six shield volcanoes built of highly fluid, peralkaline trachytic (60-67% SiO_2 , 10-12% alkalis, 1% MgO) lavas and ash flows formed 2 to 6 m.y. ago in the northern rift valley of Kenya (Webb and Weaver, 1976). The lavas and tuffs were erupted along the rift margin, forming a chain 100 km long by 30 km wide. The trachyte shields have been deeply dissected, cut by fissures, down dropped into the developing rift valley, and partially covered by younger extrusives; complete morphometric data is thus unobtainable. For the largest shield, Ribkwo, D_s is ~45 km, S_h is ~750 m, and the partially preserved caldera has a maximum diameter of 5.5 km. The volcanoes Kafkandal and Nasaken have indistinct calderas with D_c values of ~7 km and 5 km respectively; calderas are not clearly defined on the other shields.

Both the central source zones and the flanks of the shields were built up simultaneously, with highly fluid lavas and welded tuffs forming the low angle (5°) slopes of the flanks, while completely degassed magma erupted as short, thick lavas, plugs and domes at the sources. Webb and Weaver (1976) suggested that the very mobile behavior of the flank deposits (despite the high SiO_2 content) was due to a high volatile content and the peralkalinity of the magma.

Ignimbrite shields. When ignimbrites are the major products of an eruption from a central vent a broad, low shield is constructed. Although later eruptions of less fluid products often bury and disguise the ignimbrite shield, this has not happened in many places, including Japan, Tibesti and Italy. Matumoto (1943) described four ignimbrite shields (Aso, Aira, Ibusuki (or Ata), and Kikai) in the Kyusyu region of Japan (Kuno, 1962). Each shield was formed by immense eruptions of ignimbrite sheets of andesitic composition, and minor flows of lava. Ignimbrite- and pumice-filled valleys extend more than 100 km from Aso and the approximately circular shield of continuous deposits is 45 km wide. The Aira shield covers an even larger area (~100x120 km) and has an average slope less than 1° . Following the ignimbrite eruptions the summit of each shield collapsed forming large elongated calderas: Aso, 24x18 km; Aira, 23x17 km; Ibusuki, 25x12 km; Kikai, 23x16 km. The complicated outlines and circular embayments of the calderas suggest there were many overlapping collapse centers, each presumably related to a major ignimbrite eruption. Following the ignimbrite caldera phases

the margins and floors of each shield were colonized by strato-volcanoes, domes, and cones of less gas-rich magmas.

Giant ignimbrite shields have also been described from the Tibesti Mountains of Africa (Vincent, 1963). Voon, Yirrigue, and Pre-Touside are ignimbrite shields ("shield nappes" of Vincent) with $D_s \approx 100$ km, gentle slopes of 2° - 3° , and summit calderas with D_c values of 18×15 km, 13×14 km and 13 km, respectively. Tibesti also contains normal basaltic shields as well as numerous andesitic shield volcanoes (e.g., Toon, Yega, Oyoye, etc.) with D_s values of 40-60 km and $S_s \approx 10^\circ$ - 15° . These shields are composed of lavas, not ignimbrites, and have been uparched by acidic intrusions.

An alignment of four ignimbrite shields occurs in the Roman Volcanic Province of Italy. Ignimbrite deposits of one of the shields, Vulsini, have been studied in detail by Sparks (1975). A low shield with slopes of 1° - $1\frac{1}{2}^\circ$ on the outer flanks, Vulsini has a basal diameter of 50-55 km, and two adjacent summit calderas, Latera ($^\circ$ km) and Bolsena (11×13 km). Seventy percent of the volcano is composed of ignimbrites with subsidiary airfall and surge deposits. Near the caldera rims lavas and pyroclastic fall deposits dominate and the slope increases to 2° - 8° . The calderas collapsed following major ignimbrite eruptions, and minor parasitic cones have since formed on the shield's flanks and on the rims and within the calderas. There is a wide range in composition for the Vulsini ignimbrites (SiO_2 ranges from 50-62%) suggesting that magma composition is not the most important factor in determining volcano morphology.

Comparisons of basaltic and felsic shields

Vincent (1963) appears to have been the first investigator to notice the gross similarities of basaltic and felsic shields. He pointed out that the ignimbrite shields of Tibesti share the following characteristics with the classic Hawaiian shields: (1) comparable basal diameters; (2) convex profiles; (3) central calderas; and (4) radial and parallel dikes.

Vincent believed that the two types differed in that the ignimbrite shields had lower slopes than basaltic shields (2° - 3° vs. 5° - 10°), and consequently had considerably smaller volumes. Vincent used the massive Hawaiian shields in his comparison; Whitford-Stark (1975) has since documented a continuum in sizes of basaltic shields, from small "Scutulum types" or "lava cones," through Icelandic shields, to the giant shields of Reunion, Galapagos, and Hawaii. Thus, the ignimbrite and other felsic shields have volumes comparable to all but the largest terrestrial shield volcanoes. Additionally, the increased sample of basaltic shields illustrates that slopes for both basaltic and felsic shields commonly range from 1° - 10° .

Details of lava stratigraphy on basaltic shields -- flows, lava tubes and channels, collapse pots -- have some analogs on the flanks of felsic shields. Although Suswa is one of the few felsic shields with an extensive system of lava tubes and collapse pots (Williams, 1963), many others have pronounced lava flow fronts. Ignimbrites commonly form sheets with width/length ratios much greater than for basaltic lava flows, but no quantitative data is available on this topic. Smaller volume pyroclastic flows occasionally develop leveed channels and lobe-like flow fronts (Sparks, 1975), but larger flows are more sheet-like and apparently lack channels.

Major differences between the two types of shields occur in caldera size and associated volcanics. The calderas in the largest basaltic shields are only 3-5 km wide, equalling about 1-2% of their shields' basal diameters; for small basaltic shields, D_c is ~5% of D_s . In contrast, calderas of the felsic shields discussed here are much larger with D_c averaging ~32% of D_s . Additionally, the felsic calderas are often severely modified by post-collapse eruptions of viscous flows, domes, and stratovolcanoes. On the other hand, basaltic shields generally have only minor post-collapse constructs, and the original form of the shield is readily discernible. Although the shields discussed are less than a few million years old, uneven rates of denudation result in differing degrees of preservation. Basaltic and other shields built of lava flows can maintain their original morphologies longer than ignimbrite shields which are readily eroded.

Shields on the Moon, Mars, and Venus

Lunar domes are often interpreted as shield volcanoes and Head and Gifford (1977) compare them to the basaltic shields of Iceland and the Snake River Plains. The similar albedo and texture of lunar domes and the basaltic lunar mare support this interpretation. The fact that for lunar domes $D_c \approx 20\% D_s$, compared to ~2-5% for terrestrial basaltic shields and 32% for ignimbritic shields, does not detract from the interpretation of the domes as basaltic shields. A simple analysis of summit crater collapse (Wood, in prep.) illustrates that the low lunar gravity and high lava density may account for the larger craters on lunar domes.

Olympus Mons and the other giant conical mountains of Mars are generally considered to be basaltic shield volcanoes (Carr, 1973), although King and Riehle (1974) argued that the lower slopes may be composed of ash-flow tuffs. The slopes and height-diameter ratios for the martian shields are not diagnostic of the petrology or eruption mechanism for the mountains, and the ratio of D_c/D_s (15-20%) is intermediate between values for terrestrial basaltic and felsic shields.

Radar altimetry studies of Venus have revealed a 700 km wide, 10 km high mountain (called Beta) with a summit depression 90 km wide that has been interpreted as a shield volcano (Eberhart, 1977). A second volcano, 350-450 km wide with a summit crater 80 km across has also been discovered. With a slope of $\sim 20^\circ$ Beta is somewhat displaced from the D_S - H_S trend extrapolated from terrestrial shields. Both Beta and the second Venusian volcano have D_C/D_S values of 13-20%, similar to the shields of Mars.

Conclusions

The morphology of basaltic shield volcanoes is mimicked by rocks of more silicic compositions erupted either as lava flows, pyroclasts, or a mixture of the two. Morphologic details for more than a dozen shields formed of felsic lavas and pyroclastics are summarized. From this review it is apparent that the major difference in morphologies of basaltic and felsic shields concern their calderas. Felsic calderas are substantially larger and more heavily modified by post-collapse volcanism than are basaltic calderas. Proposed shields on the Moon have morphologies consistent with basaltic lava flow origins, but the shields on Mars and Venus have characteristics intermediate between terrestrial basaltic and felsic shields.

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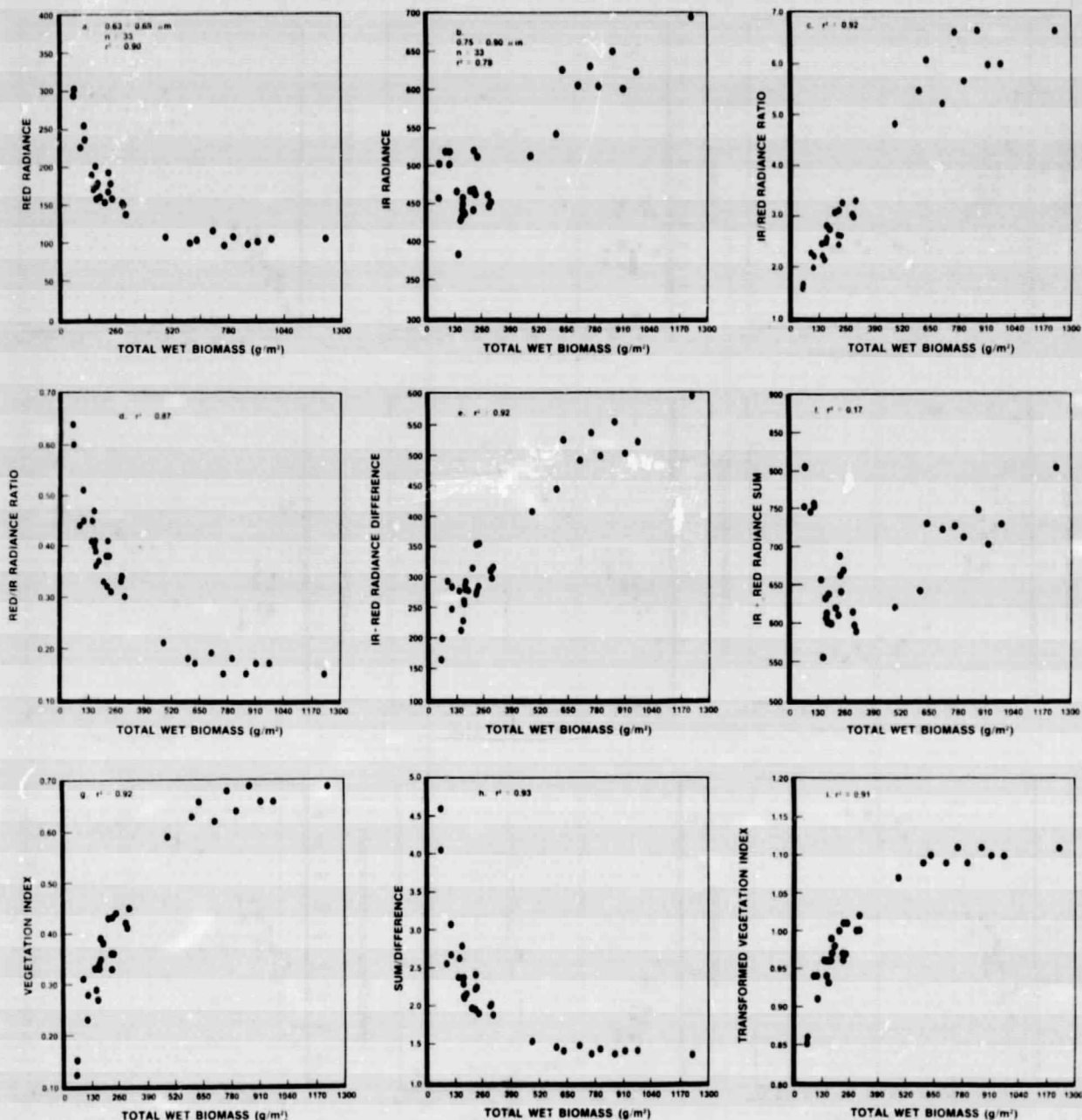


Figure 9. The nine radiance variables plotted against the total wet biomass for the 33 plots sampled in June, 1972 with the 0.70 - 1.60 μ m grating and the ir detector. The ir data used in the radiance transformations and presented in (B) is from 0.75 - 0.90 μ m. (A) red radiance, (B) ir radiance, (C) ir/red ratio, (D) red/ir ratio, (E) ir - red radiance difference, (F) ir + red radiance sum, (G) vegetation index, (H) sum/difference, and (I) transformed vegetation index. Refer to Tables 3 and 6 for tabular results.

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